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A monitoring design for the Atlantic meridional overturning circulation

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[1] Current hydrographic data can provide snapshots but no continuous timeseries of the meridional overturning circulation (MOC). Using output from two eddy-permitting numerical ocean models we test the feasibility of a monitoring system for the MOC in the North Atlantic. The results suggest that a relatively simple arrangement, using moorings placed across a longitude-depth section and the zonal wind stress, is able to capture most of the MOC strength and vertical structure as a function of time. Being closely related to the transport of energy to the North Atlantic, measuring the MOC would open the prospect of having continuous information about a key element of northern hemisphere climate. *INDEX TERMS:* 4532 Oceanography: Physical: General Circulation; 4594 Oceanography: Physical: Instruments and techniques; 4576 Oceanography: Physical: Western boundary currents. **Citation:** Hirschi, J., J. Baehr, J. Marotzke, J. Stark, S. Cunningham, and J.-O. Beismann, A monitoring design for the Atlantic meridional overturning circulation, *Geophys. Res. Lett.*, *30*(7), 1413, doi:10.1029/2002GL016776, 2003.

1. Introduction

[2] Surface currents (e.g., the Gulf Stream) bring warm and salty surface waters to the high latitudes of the North Atlantic, where cooling increases sea surface density. Eventually, surface waters sink and flow southward as cold deep currents, thus contributing to the thermohaline circulation (THC) [*Dickson and Brown*, 1994]. The THC is the dominant part of the meridional overturning circulation (MOC), which also encompasses meridional mass transports that are wind-driven. The MOC results in a net northward heat transport in the Atlantic of about 1 PW (10^{15}W) at 25°N [*Hall and Bryden*, 1982]. This is approximately half of the total northward energy transport in the North Atlantic area, the other half being provided by atmospheric transports [*Trenberth and Solomon*, 1994]. Numerical simulations suggest that under greenhouse gas increase the MOC may weaken or break down [*Cubasch et al.*, 2001].

[3] Hydrographic data allow snapshot estimates of the MOC and related transports [*Hall and Bryden*, 1982; *Ganachaud and Wunsch*, 2000], but monitoring the MOC continuously in this way would require an unrealistically high number of sections. The Florida Strait transport can be measured continuously using the voltage induced in submerged telephone cables; at 27°N this method has been used for over 20 years to monitor the annual cycle as well as the long term transport of the Florida Current [*Baringer and Larsen*, 2001]. For an estimate of the MOC this is not sufficient since additional knowledge about the southward return flow, including its vertical structure, is required. The vertical flow structure for past meridional transports can be deduced from δ^{18}O of calcite in foraminifera [*Lynch-Stieglitz et al.*, 1999] but this approach is not suited for monitoring the current MOC.

[4] Here we present a detailed concept for continuously observing the Atlantic MOC. This implies an identification of the dominant terms in the force balance governing the MOC, which so far has only been done in idealized models [*Marotzke*, 1997]. We test our design for a monitoring array in eddy-permitting numerical models, which provide comparisons between the "ideal case" (complete knowledge of meridional velocity field) and "feasible cases" where the MOC is estimated from zonal density and wind information only.

2. Measuring System

[5] The MOC is decomposed into contributions related to the zonal wind stress acting at the ocean surface and to zonal density differences, respectively [*Lee and Marotzke*, 1998]. Winds can be inferred from satellite data, atmospheric analyses and ship measurements, whereas vertical density profiles of sea water, obtained by moored profilers [*Morrison et al.*, 2001] or by sensors at fixed depths, can provide zonal density gradients. We focus our investigation on latitudes around 26°N. This is one of the best measured ocean areas and it is close to the meridional heat transport maximum [*Trenberth and Solomon*, 1994], thus capturing the entire ocean heat transport convergence into the North Atlantic. The availability of cable measurements in the Florida Strait means that the total mass transport there can be assumed to be known at all times. Every three months the meridional velocity field v_F across the Florida Strait is taken from the numerical models. Between these updates v_F is scaled by a constant so that its integral matches the actual total transport in the Florida Strait. To mimic inaccuracies in the cable based observations, a noise with standard deviation 1 Sv is added to the modeled Florida Strait transport [*Baringer and Larsen*, 2001].

[6] A schematic of a monitoring arrangement based on density and wind contributions as well as on the mass transport through the Florida Strait is shown in Figure 1. The meridional transport east of the Florida Strait is determined using vertical density profiles and the surface zonal wind stress τ_x. Transports related to the wind stress

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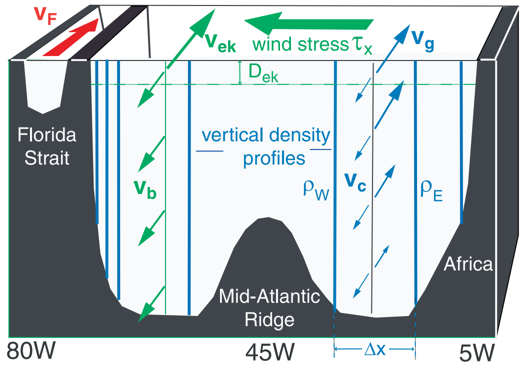

Figure 1. Schematic of an MOC observing system across 26°N in the North Atlantic. The MOC is estimated from the zonal wind stress and from vertical density profiles taken at different longitudes across the basin. The transport across the Florida Strait is assumed to be known (v_F). Knowing the wind stress allows the calculation of an Ekman velocity v_{ek}. A constant return flow with velocity v_b is assumed over the whole water column in order to ensure zero net meridional transport related to the wind stress. East of the Florida Strait a depth-dependent velocity field v_g is estimated using the thermal wind relation between adjacent profiles of density. The meridional transport associated with v_g and v_F is compensated by a spatially constant velocity correction v_c, in order to ensure zero net meridional transport for the thermal wind contribution.

(Ekman transports) can be estimated from theory once the wind field is known [*Gill*, 1982]. Assuming a depth D_{ek} for the surface Ekman layer (typically 50–100 m) a meridional velocity can be derived:

$$v_{ek} = -1/(\rho^* f A) \int_o^L \tau_x dx,$$

where L is the zonal basin width, ρ^* a reference density, f the Coriolis parameter and A the cross section of the Ekman layer. This velocity is compensated by a depth independent (barotropic) return flow v_b in order to avoid a net meridional mass transport. Note that the time mean of the zonal wind stress also affects density gradients and thus generates a depth-dependent (baroclinic) geostrophic flow [*McCreary and Lu*, 1994]. However, since the mean value of the zonal wind stress at 26°N is small compared to its variability, we assume a barotropic return flow.

[7] Density is observed through vertical profiles placed at different longitudes across the Atlantic basin (vertical lines east of the Florida Strait, Figure 1). Between adjacent vertical profiles (e.g. ρ_W, ρ_E in Figure 1) a depth dependent velocity is obtained based on the thermal wind balance [*Marotzke et al.*, 1999]:

$$v_g(z) = -g/(\Delta x f \rho^*) \int_{-H}^{z} (\rho_W - \rho_E) dz.$$

The zonal distance between the profiles is Δx, H is the ocean depth and g is the acceleration due to gravity. Using v_g as well as the velocity field v_F in the Florida Strait, a depth integrated zonal mass transport can be calculated. As for the wind contribution, the net meridional transport is corrected to zero by a spatially (but not temporally) constant correction v_c which does not distort the vertical shear of the geostrophic flow v_g obtained with the thermal wind relation [*Hall and Bryden*, 1982]. Note that the results for the MOC reconstruction would be the same if wind and density contributions were first added and then corrected by one single correction to close the mass balance. We use two corrections v_b and v_c so that we can give an estimate of the relative contribution of each part of the measuring system.

3. Models

[8] The numerical models in which the proposed monitoring arrangement is "deployed" are the 1/4° version of the global model OCCAM (Ocean Circulation and Climate Advanced Modelling Project) [*Webb*, 1996] and the 1/3° version of the Atlantic model FLAME (Family of Linked Atlantic Model Experiments) [*Dengg et al.*, 1999]. Sea surface temperatures and salinities are restored to monthly climatological values in OCCAM [*Levitus and Boyer*, 1994; *Levitus et al.*, 1994] and the wind stress is based on the ECMWF (European Center for Medium-Range Weather Forecast) climatology. In FLAME sea surface salinity is restored to climatological values [*Levitus et al.*, 1994]. Net heat flux and wind stress data are based on a 3-year climatology of ECMWF analyses and on monthly mean anomalies.

[9] The reason for using these two models lies in the different strengths of the simulated MOC: In OCCAM the meridional circulation is weak with average values of about 6 and 10 Sv (1 Sv = 10^6 m^3/s) at 45°N and 25°N, respectively, whereas much stronger values are found at the same latitudes in FLAME with average values of about 20 Sv and 16 Sv. Using both models increases our confidence that changes of the circulation can be monitored over a large range of different MOC strengths.

4. Transport Estimates

[10] Transport reconstructions are summarized in Figure 2 for OCCAM and FLAME. In both models the monitoring array for the thermal wind contribution consists of 9 "moorings": four profiles are placed along the western boundary east of the Florida Strait and three are placed along the eastern boundary. Furthermore, two profiles are located west and east of the Mid-Atlantic ridge (panel a). The decision to use 9 profiles was the result of tests with arrays using larger and smaller numbers of profiles. While the quality of our results remains unchanged for higher numbers of profiles, using fewer profiles along the eastern and western margins of the basin degrades the vertical structure of the estimated meridional flow.

[11] Maximum values of the MOC and of its reconstruction (panels b, c) show that the strong seasonal variability is well represented in the reconstruction. The bias of about 2 Sv for the mean value in FLAME is not fully understood yet. One explanation could be the choice of a spatially constant correction v_c for the thermal wind contribution. Other corrections using different constants between each pair of adjacent profiles also preserve the vertical shear

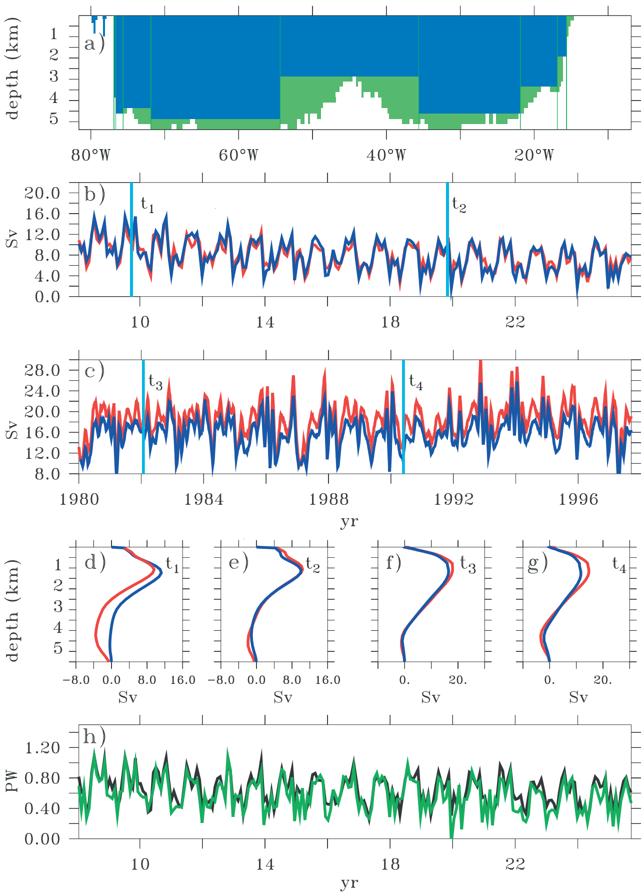

Figure 2. Transport reconstructions based on 9 vertical profiles of density and Ekman contributions. **a**, Distribution of vertical density profiles in OCCAM (vertical green lines). The blue shading indicates where the vertical profiles provide an estimate of the velocity shear. In the green areas (bottom triangles) the vertical shear cannot be estimated from adjacent profiles. **b** and **c**, Timeseries (18 years) for reconstructed maximum meridional overturning (blue) and maximum overturning using the full model velocities (red) at latitudes of 26.5°N (OCCAM, panel b) and 26°N (FLAME, panel c). **d–g**, Vertical flow patterns for the MOC (red) and its estimate (blue). Snapshots are shown at times t_1, t_2, t_3, and t_4 for OCCAM (panels d, e) and FLAME (panels f, g). **h**, Net meridional heat transport at 26.5°N in OCCAM (black) and heat transport calculated from estimated mass transport (green).

obtained from the thermal wind relation but can modify the MOC structure and value.

[12] The vertical structure of MOC and of the estimated transport at times t_1, t_2, t_3, and t_4 (vertical lines in b and c) is shown in panels d-g. In both models, the maximum value for the MOC and the reconstruction is found at a depth of about 1000 m. Below, the strength of the return flow is reasonably well reconstructed in both models, although at times the near-bottom circulation is underestimated in OCCAM (panel d).

[13] Panel h shows the net heat transport across 26.5°N based on meridional velocities obtained from the "monitoring array" (green) and the model velocities (black). For the reconstructed heat transport the temperature field across the Atlantic section is assumed to be known at the start of the timeseries and kept unchanged afterwards. This reflects that changes in the velocity field are rapid compared to variations of the temperature field [*Jayne and Marotzke*, 2001]. The estimated heat transport follows the model heat transport very closely, both in the time mean (0.7 PW) and the variability (1 PW peak-to-peak).

[14] During the 18 years of integration there is a gradual weakening of the MOC in OCCAM, masked by the strong seasonal variability in Figure 2. In Figure 3 the decrease of the MOC becomes apparent after the timeseries have been smoothed by a two year running mean. The maximum MOC strength decreases from more than 10 Sv to less than 7 Sv, a reduction that is captured well by the simulated observing array. As shown by the light blue envelope, adding uncertainties in the Florida Strait transport and the density field does not affect the detectability of the gradual decrease in the MOC strength.

[15] Considering the Ekman and thermal wind contributions separately shows that the contribution obtained from vertical profiles of density dominates the MOC (Figure 4). The thermal wind contribution exhibits a clear seasonal variability in both models. For OCCAM it is obvious that the decrease of the meridional mass transport discussed previously is found in the thermal wind contribution, thus indicating a gradual change in the east-west density differences. With a mean value of 1–2 Sv in OCCAM and FLAME, the wind contribution adds only little to the mean total transport. However, it exhibits a pronounced variability with peak-to-peak amplitudes up to 8 Sv in both models. This shows that a reliable measurement system must include both density and wind information.

[16] Note that in both OCCAM and FLAME the bottom velocities are very small at the latitude of 26.5 N. Therefore, we assume $v_g = 0$ at the ocean bottom. Major deviations of the method used here are expected in cases where high velocities are found at the ocean bottom. In that case a substantial barotropic contribution would not be captured by the velocity v_g. In the real Atlantic this is most likely to be

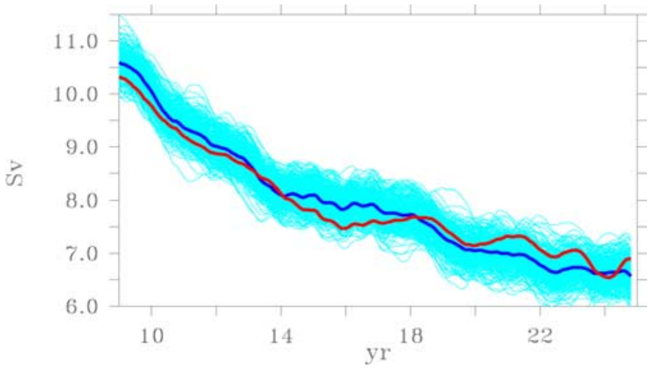

Figure 3. Evolution of maximum values for 300 estimates of the MOC (light blue lines) obtained with noise (standard deviation 1 Sv) added to the mass transport through Florida Strait and to the density field (standard deviation 0.01 kg/m^3) in OCCAM. The values are smoothed by a two year running mean. The solid blue line indicates the average of all reconstructions whereas the red line shows the maximum MOC values based on the full model information.

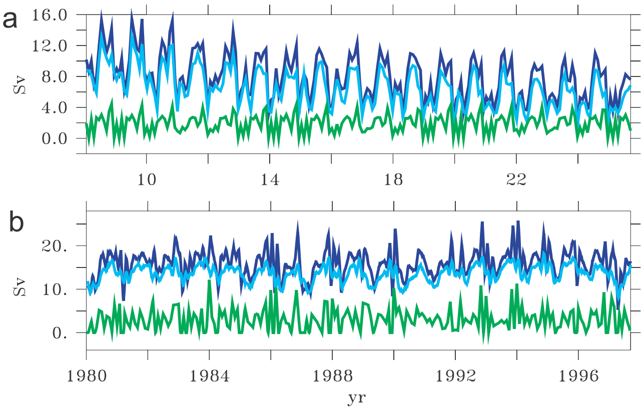

Figure 4. Ekman (green) and thermal wind (light blue) components in OCCAM (a) and FLAME (b) obtained from the zonal wind stress and 9 vertical density profiles, respectively. The sum of both components is depicted in blue.

the case along the western margin, where the deep western boundary current can hit the continental slope [*Lee et al.*, 1996]. Bottom velocities from direct measurements or bottom pressure recorders should be used as a reference for the calculation of v_g. As before, the mass balance could be ensured by using a spatially constant correction v_c.

5. Conclusions

[17] For the first time simple force balances have been used to estimate the MOC in eddy-permitting models with realistic topography and forcing. Our results suggest that continuous monitoring of the MOC is possible with existing technologies. The acquisition of data needed to recover most long and short term variability could be done at comparatively low costs.

[18] However, knowing the MOC and heat transport is only the first step. If a change is detected the proposed measuring system can only provide rudimentary information about the underlying mechanisms (e.g., density anomalies seen in the profiles). The monitoring system should be used in combination with measurements from other latitudes, thus allowing an estimation of signal propagation in the ocean. To further increase the confidence in the method, tests with models using higher resolutions would be very useful. In the near future a continuous monitoring system similar to the one described in this study will be deployed in the framework of the U.K. Rapid Climate Change programme [*National Environmental Research Center*, 2003]. The design study presented here indicates that the data acquired by such a system is able to provide crucial information about the North Atlantic MOC and its variability.

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